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THE MECHANICS OF GLACIERS. I.[1]

1. *Law of flow.*—If we consider the ice flowing through two cross-sections of a glacier, it is evident that for a glacier in equilibrium, that is, one neither increasing nor diminishing in size, the quantity flowing through the lower section must equal that flowing through the upper one diminished by the quantity melted between them, if we are in the region of melting (the dissipator); or increased by the quantity of ice accumulated between the sections, if we are in the region of accumulation (the reservoir). Keeping the upper section at the névé-line and moving the other one down the glacier, the melting area between the sections increases and therefore the flow through the lower section becomes less. Now placing the lower section at the névé-line and moving the upper we see that the amount of accumulation between the sections increases, and consequently the flow through the upper section diminishes, as we move it higher up the reservoir. If we keep our two sections at a fixed distance apart, and move them to different parts of the glacier, the difference of the flow through the sections must increase as we go down the dissipator, because the rate of melting increases, and must increase as we go up the reservoir because the accumulation increases. We can say then that *the greatest flow occurs through a section at the névé-line and diminishes as we go up or down the glacier from there;* and, moreover, that *the rate of diminution of the flow becomes greater the further we go from the névé-line.* This is the *law of flow*, and is perfectly general; it depends only on the law of the indestructibility of matter, and the observed facts that the ice of glaciers moves, and that there is accumulation above, and dissipation below, the névé-line.

The steeper the dissipator's slope the more rapidly will the rate of melting increase as we go down the glacier, and the more rapidly will the flow diminish; similarly, the steeper the reservoir the more rapidly in general will the flow diminish as we ascend.

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[1] Read before the Geological Society of America, at the Philadelphia Meeting, December, 1895.

It has been stated as an empirical rule that the velocity of the ice of glaciers lying in beds of uniform slope is greatest in the neighborhood of the névé-line and diminishes as we leave it going up or down the glacier. This rule is, like most empirical rules, subject to exception, and may be stretched beyond its limit, as when Professor Heim (Gletscherkunde, p. 160) attempts to account for the increased velocity near their ends of some Greenland glaciers which break off in bergs, by saying that we have here to do with only the upper part of the glacier; whereas, in reality, the ends are some distance below the névé-line. The general law that the flow is less below, than at, the névé-line must hold; this flow equals the product of the average velocity by the sectional area by the effective density. The more rapid surface velocity is permitted by the formation of immense crevasses which reduce the effective density of the upper 200 feet of the ice by perhaps a half. The real and sufficient cause of this increase in velocity is the lack of support in front, which Professor Heim also describes. The flow near the end of such glaciers is not as large as appears at first sight; it is only the upper part of the ice that has such an abnormal velocity; the lower part, being supported by the pressure of the water, must have a velocity not differing very greatly from what it would have if the glacier completed its course and ended as an ordinary alpine glacier. Indeed, as the increasing velocity at the surface is only permitted by the opening of immense crevasses, so an increasing velocity in the middle and under part of the glacier would also be accompanied by similar openings; and the whole body of the ice would be so torn by these great fissures, that comparatively thin sheets would be broken off from the glacier's end, and large icebergs could not be formed. The drag of the upper layers over the lower ones, would, in glaciers whose ends are floating, have to be entirely balanced by the cohesion of the lower ice-layers, as there is no friction against its bed.

The empirical rule does not cover the case of a glacier like the Mer de Glace, where the bed becomes steeper near its end, causing a more rapid motion. But the law of flow must hold; here again the effective density of the ice is reduced by the opening of wide crevasses, and the sectional area is undoubtedly less than it would be if the slope of the glacier's bed continued uniform.

In the case of glaciers with beds of uniform slope, the velocity and flow increase and decrease together, though not in the same proportion.[1] If we had a glacier of indefinite length and of uniform section, it is evident that the direction of flow would be parallel with the slope, and the velocity along any line parallel with the glacier's axis would not vary as we move along the direction of flow. To calculate the forces acting on a section of such a glacier we have only to consider the weight of the section, the viscosity of the ice and the friction against the bed. The motion consists of two parts: (1) the sliding on the bed, of which we have hardly any knowledge; (2) the shearing of the section by virtue of which $A\ B$ (Fig. 1) would be deformed into $A'B'$. Let us call the velocity of a point under such conditions the *normal velocity* corresponding to that particular form and size of the cross-section.

A glacier of uniform section could not exist if there was any melting; for we have seen that in this case the flow would become smaller as we go down the glacier, and with uniform section this would require a decreasing velocity, which, however, is not admissible; for, from the supposed uniformity of the glacier, every section would be subject to the same forces and would move with the same velocity. Therefore, the slope of the glacier being uniform, wherever there is melting the cross-section must change so as to produce a smaller flow as we descend the glacier. For glaciers which fill narrow valleys this simply means a diminution in both the breadth and thickness of the glacier; but for those which spread out, like the Davidson glacier in Alaska, the flow is diminished by the smaller thickness, in spite of the increased breadth.

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[1] The remainder of this section and all of the next are based on the theory that the ice of glaciers acts in general like a viscous substance, and that the sole cause of the motion is the weight of the ice itself. I am not discussing the physical nature of ice which gives it this viscous property. Similar conclusions might be reached on other theories. The rest of the paper is only dependent on the observed phenomena of glaciers.

We may infer a similar thinning of the glacier above the névé-line; indeed, with many glaciers, the great breadth of the reservoir, notwithstanding the smaller density of the ice, requires a very small depth in order that the flow should not be too great.

If we look upon the ice-sheet, which covered North America, diagrammatically as radiating from a center, and suppose that the annual accumulation and melting per unit surface were proportional respectively to the distance inside and outside of the névé-line, we find, in order that total melting and accumulation should be equal, that the névé-line would have been at about two-thirds of the distance from the center to the circumference of the circle. It is probable, however, that the melting was larger at the edge, and the accumulation less at the center than this, and, therefore, that the névé-line was nearer the circumference. The greatest flow was of course through a section through the névé-line, and the greatest velocity must have been somewhere in the same neighborhood.

Pressure.—Let us see if the different parts of a glacier, gradually thinning out to its ends, can have the *normal velocities* due to their cross-sections. We know next to nothing about the sliding of the ice over its bed, but we are safe in assuming that the normal velocity of sliding would be no greater in a thin than in a thick section.[1]

Consider then that the bottom velocity of the region between the two sections *AB* and *CD* of the dissipator is uniform (Fig. 1). For the sake of simplicity we will consider the thickness of the glacier small in comparison to its breadth. After a certain time the sections would be deformed to $A'B'$ and $C'D'$, if the ice moved everywhere with its normal velocity. The slope of the deformed section at any point would depend only on the weight of the ice above that point, *i. e.*, on its depth below the surface of the glacier, and therefore the form of the section $C'D'$ would be exactly the same as that of the upper part of $A'B'$. As the slope of $E'B'$ is greater than that of $F'B'$, it is evident that more ice would enter through EB than would leave through FD in the same time, EF' being parallel to the bed of the glacier; but this is impossible under the supposition of normal flow, for we have only considered a part of the glacier below the surface, where no melting takes place.[1] The tendency of the thicker sections to move faster than the thinner ones, at the same distance from the bed, causes a forward pressure on the latter, increasing their velocity, and a backward pressure on the former, diminishing their velocity. The ice near the lower end of the glacier is thus under a pressure greater than the *normal pressure*[2]

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[1] The only experiments bearing on this point are those of HOPKINS (*Phil. Mag.*, London, 1845, Vol. XXVI, pp. 3–6). By means of a wooden frame he held together a

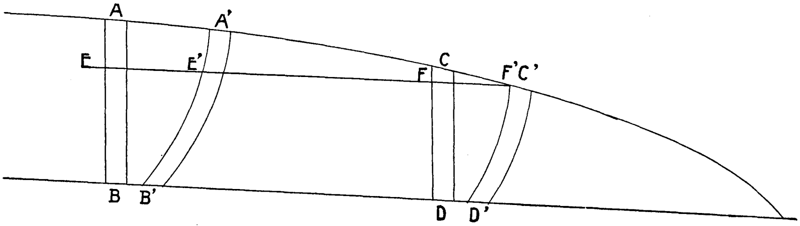

FIG. 1. Displacement of sections by flow.

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number of pieces of ice on a slightly inclined rough sandstone slab, and found that the mass slowly slid down the slope with a uniform velocity approximately proportional to the pressure against the slab and to the angle of inclination, for angles between 1° and 10°. When the sandstone was smoothed but not polished a motion was observed at an inclination of forty minutes. The ordinary laws of the friction between solid bodies do not apply at all to the forces between a glacier and its bed.

[1] We here neglect the melting at the lower surface of the glacier which is so small as not to alter the argument.

[2] The normal pressure is the pressure which would exist if the glacier were of infinite length and uniform section.

and would therefore have a tendency to rise, would be squeezed up, so to speak.

We cannot reason very definitely about the pressure above the névé-line, for the much increased breadth of the reservoir, the smaller density of the ice and the possibly greater slope of the bed may result in either pressure or tension according to their relations to each other and to the thickness of the glacier.

The idea of a pressure on the lower reaches of a glacier due to the ice above has been generally held by glacialists, but I think without satisfactory reasons. It seems to have been inferred generally from the diminishing velocity of the lower parts of the glacier ; but it is not due to this, for it is quite conceivable that by an increasing slope, and suitable melting, every section might have its normal velocity ; there would be no effective pressure upon the upper regions, and there might still be a diminishing surface velocity.

Indeed, the extra pressure is not due simply to diminishing flow, but to the fact that the flow diminishes more rapidly than the cross-section. This may be stated in algebraic language by saying that the normal flow is not proportional to the linear dimensions of the section; it is not even proportional to the square of the linear dimensions, but probably to the third power.

In the dissipator of a circular ice-sheet, or of a spreading glacier, successive sections increase in breadth with a more rapid diminution of thickness than in the case of a linear glacier. This causes a considerable pressure of the rear on the forward sections, resulting in a more rapid radial movement of the ice than would occur without this pressure and in the opening of radial crevasses. It is not unlikely that such crevasses determined the positions of the eskers formed during the glacial epoch.

Direction of flow — Stratification.— In order that the general volume of the glacier should be preserved we must have below the névé-line, where there is melting, a component of the motion towards the surface, and this component is strongest where the melting is greatest, *i. e.*, it gets larger as we descend the dissipator. On account of the gradual thinning of the glacier this condition would be satisfied if the motion were everywhere parallel to the glacier's bed, but we have already seen that on account of the pressure from behind there is really a small movement of the ice away from the bed. Above the névé-line this is reversed. Here there is accumulation; and, unless the reservoir continually increases in thickness, the motion must have a component downward into the glacier, and this component must be greater where the accumulation is greater, *i. e.*, in general as we ascend the reservoir. We can thus draw on the surface of the glacier the approximate directions of motion, and beginning at the névé-line, where, as there is neither melting nor accumulation, the motion must be parallel to the surface, we can connect these lines and have a diagram of the lines of flow in the whole body of the glacier (Fig. 2). The slight melting of the glacier in contact with its bed would produce a small component of the motion towards the bottom, and in the dissipator also towards the sides; but in the reservoir the accumulation results in a movement away from the sides. In picturing to ourselves the relative velocities of points at different distances from the surface, it is best to consider points lying in a section drawn everywhere at right angles to the lines of flow. The velocity of points in such a section will diminish as we approach the glacier's bed, on account of the friction which acts on the ice there. We can also show the positions assumed by the strata of annual accumulations. Perhaps the best way to do this is to follow the successive positions of a chosen stratum. At the end of the summer the snow deposited during the previous year will end in a thin wedge at the névé-line, and its thickness will at every point equal the corresponding annual accumulation. A year later the end of this stratum, with but small loss by melting, will have been carried a short distance down the surface of the glacier; and every point of the stratum following its proper line of flow, will have progressed down the valley and into the body of the glacier, and the stratum will have reached the position *2*. Continuing in the same way we can trace in a general way the position the stratum will have in successive years; and these positions, at any one time, will correspond to those of successive annual strata.[1]

We readily see that the strata have a gentler slope than the lines of flow in the reservoir, but a steeper slope in the dissipator; and that outcrops of strata can only occur below the névé-line.

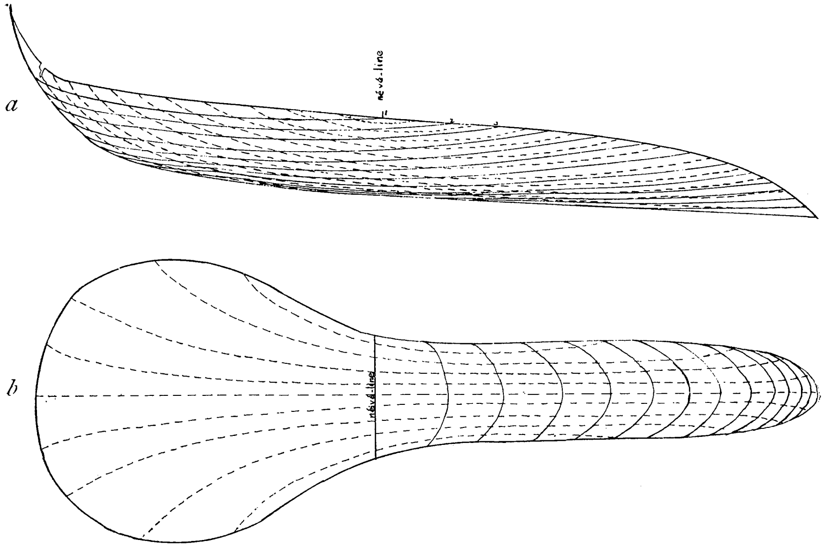

FIG. 2. Diagram of stratification (*a*), and lines of flow (*b*). Section and ground-plan.

The lower end of a new stratum is in the part of the glacier where the horizontal motion is greatest, and it never sinks far into the body of the dissipator, whereas, the upper end of the stratum originates in a region of small horizontal motion and later forms the lower layers of the glacier; the stratum, therefore, is stretched out and grows thinner on account of the differential motion, and also on account of the compression of the snow into ice. During the first part of its course the angle which it makes with the bed of the glacier steadily increases; but later diminishes until probably it at last becomes nearly parallel with the glacier's bed.

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[1] Agassiz (Syst. Glac., pp. 273–276) very nearly reached this conception of the disposition of strata and the direction of flow. Reverend O. Fisher (Phil. Mag., London, 1879, Vol. VII, pp. 381–393) presents somewhat similar ideas in connection with the stratification of Antartic ice. The theory presented above was developed before I was familiar with the latter's work.

It is to be noticed that the largest accumulation occurs under the mountain cliffs where the glacier receives, in the form of avalanches, much of the snow that falls on the slopes above. It is here then that the vertical component of motion is greatest, and it is also in this neighborhood that we find the *bergschrund;* which marks the line where the more rapidly moving ice below pulls away from the more slowly moving ice above. Reverend Coutts Trotter[1] has made the important suggestion that the *bergschrund* may mark the line above which the mean annual temperature is below freezing, and therefore the ice above it is frozen to the mountain, whereas, the ice below slides on its bed in addition to its shearing motion. Unfortunately, there is, so far as I know, no experimental evidence bearing on this point. Without disputing Mr. Trotter's idea, we see that the general theory leads us to expect a considerable change in motion in this part of the glacier; and when we recall that the surface slope diminishes rapidly below the *bergschrund*, we are driven to infer a rapid increase in the thickness of the glacier, and at least a large factor in the formation of the *bergschrund* becomes clear; for above it there is a thin coating of ice on the mountain slope, and below a much thicker mass of ice resting on an almost equally steep bed; and this would certainly result in a marked difference in velocity of the two parts.

Accurate observations on the direction of motion and on the stratification are few. Désor found at the side of the Unteraar glacier a motion of the ice having components down the valley, towards the side, and upwards.[2] Agassiz found by a line of stakes across the same glacier, whose positions and levels had been carefully determined, that during the winter the ice near the center of the glacier had suffered an upward displacement of 2.25 meters, and that the average upward displacement of the whole section was 1.4 meters.[1] These amounts are given after allowing for the natural thickening of the ice due to winter progression parallel to the glacier's bed, without melting. There is no apparent reason why the directions of flow in the winter and in the summer should differ; so I think these observations indicate a continuous movement of the ice toward the surface, as we should expect.

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[1] Proc. Roy. Soc., 1885, Vol. XXXVIII, pp. 92–108. In this article Mr. Trotter describes some very instructive experiments showing that glacier ice will shear under very small forces if sufficient time is allowed.

[2] AGASSIZ, Système Glaciare, pp. 499–504.

Professor Pfaff found in one of the reservoirs of the Aletsch glacier, where the surface slope was 9°, that the direction of motion made an angle of about 40° with the horizontal. This is certainly a high angle and Professor Heim does not think the method used sufficiently free of error to establish these "somewhat astonishing results."[2] Here again, however, the observations are in harmony with theory. It is unfortunate that so little has been done to determine the vertical component of the motion.

The correct observation of stratification is very difficult. It is next to impossible to determine the surfaces of separation of successive strata in the dissipator unless we find them marked by a thin layer of débris. Aggassiz has made a drawing of several such débris layers as they were exposed in a hole in the upper part of the dissipator of the Unteraar glacier.[3] These must be true surfaces of stratification; they dip up stream at an angle of about 30°.

Professor Russell has described débris layers in the Sierra Nevada glaciers, and suggests that they separate successive strata; they occur at fairly regular intervals, dip into the glaciers, and were only seen below the névé-line.[4]

When the junction of two glaciers, which gives rise to a medial moraine, occurs in the dissipator, the moraine remains on the surface of the ice; but if the junction occurs in the reservoir the moraine is covered up by later strata of snow and reappears lower down in the dissipator; the further the junction is above the névé-line, the further is the point of reappearance of the moraine below that line. Rock material that falls on the snow from the cliffs in the cirque is carried by the flow along the under part of the glacier and reappears at the surface only near the lower end of the dissipator.

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[1] Syst. Glac., pp. 559–563.

[2] HEIM, Gletscherkunde, p. 184.

[3] Syst. Glac., p. 260. Atlas Pl. V, Fig. 15.

[4] The Glaciers of the United States. 5th An. Rep. U. S. Geol. Surv., 1883–4, pp. 316, 319. Pl. XXXVII shows the outcrops of the layers in the Mt. Dana glacier. See also HEIM, Gletscherkunde, p. 131.

We have but few observations bearing on these theoretical inferences. In Agassiz's map of the Unteraar glacier one or two small moraines are shown as beginning on the surface of the ice some distance below the névé-line. And it is not at all unlikely that what is usually called the spreading of the moraines near the glacier's end is largely due to the appearance at the surface of the débris fallen on the glacier from the cliffs surrounding the reservoir. A glance at the lines of flow (Fig. 2) will make this clear.

The Gorner, the Findelen and the Zmutt glaciers, near Zermatt, end within a few miles of each other; the first two, originating on high snow-covered land surrounded by snow ridges, have a large portion of their surfaces comparatively clean to the end; the last, whose reservoirs are dominated by precipices from which rocks are continually falling, is entirely concealed at its lower end by débris. Of the tributaries of the Muir glacier in Alaska, Dirt glacier, originating in a cirque, has two or three miles of its lower end deeply covered by débris, whereas the White glacier, nearby, whose reservoir is not surrounded by cliffs, has ordinary linear moraines.[1]

The lines of flow (Fig. 2) explain why all observers have had practically unsurmountable difficulties in following the stratification from the reservoir through the dissipator.

The parts of the strata formed at a distance from rocky slopes have very little dust blown upon them, and consequently when they reappear at the surface in the upper or middle regions of the dissipator the stratification is but slightly, if at all, indicated by dust bands. The strata should be well defined at the lower end, but the large amount of débris on the surface and in the crevasses, would make them difficult to recognize.

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[1] See Map in Studies of Muir glacier, Alaska, Nat. Geog. Mag., Vol. IV, p. 52.

The débris in the lower layers of some glaciers may have a similar origin, and may not have been picked up from the ground beneath. Indeed, in general a downward rather than an upward motion probably occurs near the bottom, and some of the material forming the ground-moraine may have been originally dust or rock fallen on that part of the reservoir's surface, near the encircling cliffs, which afterwards formed the lower layers of the glacier. For, on account of the heat derived from the earth, two feet of ice would be melted from the glacier's under surface in a hundred years, and would deposit its débris under the ice.

Form of the glacier's surface.—For equilibrium the melting of the ice at every point of the surface must exactly equal the supply; this is the necessary condition; and since the upper layers move fastest, their melting must be fastest; this is brought about by the slope of the surface being least where the motion is most rapid, thus exposing to melting a larger surface of the layer. Or we may say that the angle between the direction of motion and the surface is greatest where the motion is smallest and the melting most active. At the end of the glacier, therefore, and near the ground, we find the slope steepest; and as we pass up along the surface of the ice the motion becomes faster and the melting less and consequently the slope diminishes. This also explains the rounded plan of the glacier's end and the depressed sides. If we knew accurately the distribution of velocity and of melting we could calculate the shape of the surface for equilibrium; but without this our approximate knowledge of these quantities enables us to see that the general form of the dissipator's surface agrees with the theory.

Glaciers are continually changing in size, so that the conditions for equilibrium can be looked upon as approximately true only when the changes are small, or when an average is taken over a long period.

Form of the glacier's lower end.—Let us now give our attention to the lower end of the dissipator. We have seen that the surface slope at any point depends on the velocity and rate of melting at that point, and varies inversely with them. The general form will, therefore, depend on the differential velocity. (We are considering a portion of the glacier so small that we may suppose the melting uniform ; though this would be somewhat modified by the direction of the sun's rays.) The larger the glacier, the greater in general will be the differential motion, and the more rapidly will the slope diminish as we ascend from the end. For very small glaciers the differential motion is small and the slope remains steep. If anything should cause an abnormally rapid melting of the lower layers, the natural slope could not be retained, but the upper layers would advance over the lower ones, project and break off. Professor Chamberlin[1] in his excellent descriptions and illustrations of the glaciers of northern Greenland has introduced us to just such forms. Fig. 3 is a section taken from one of Professor Chamberlin's photographs of Bryant glacier,[2] which is a good example of the type of glacier we are now considering. It is about 140 feet thick at the end; the lower layers, for something more than half its thickness, are full of débris; the upper ones are practically clear with an occasional layer of débris. The terminus of the glacier inclines slightly forward in the dirty lower layers, above which the clear ice projects in one or two somewhat overhanging vertical steps to the upper surface of the glacier, which slopes back from this point according to the natural surface of an alpine glacier. Wherever a débris layer occurs in the clear ice there is a reëntrant groove in the cliff, and the layer above overhangs. Wherever the layer of débris is interrupted so is the groove.

All indications show a very slow motion, which is probably entirely accounted for by the small thickness of the ice.

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[1] This JOURNAL, Vols. II and III; Bull. Geol. Soc. Amer., 1895, Vol. VI, 199–220.

[2] Bull. Geol. Soc. Amer., 1895, Vol. VI, Pl. III, Fig. 2.

Professor Chamberlin has suggested that a partial cause of the vertical and overhanging ends is the greater effect of the nearly horizontal rays of the sun on the steep edges than on the sloping backs of the glaciers; this is undoubtedly a true cause; he thinks, however, it is not the whole explanation, but that there must be some special shearing at the débris-bearing layers. This seems to me unnecessary; and Professor Russell[1] has lately made clear that the presence of débris in a glacier diminishes, instead of increasing, its power to shear.

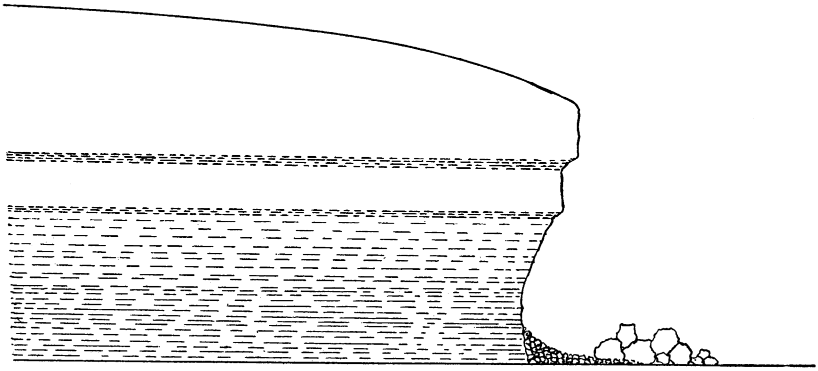

FIG. 3. Diagrammatic section of end of Bryant glacier.

The endings of these glaciers present a close analogy to the endings of tide-water glaciers. The lower layers of the latter have their extent determined by the equality between the rate of motion and the melting due to the salt water, and the upper layers project over them until they break off.[2] In these Greenland glaciers the extent of the lower layers depend on their rates of motion and of melting, and the upper layers push over them until they break off from lack of support, as is shown in the illustration referred to by the pieces of ice lying in front of the glacier. Above the vertical cliffs, thus produced, the ordinary conditions of melting prevail, and the surface of the glacier follows the ordinary alpine form. The débris is, however, an important factor in increasing the melting of the lower layers. It acts in two ways: it absorbs the sun's rays more abundantly than the ice does, and its presence reduces the effective density of the ice, so far as melting is concerned, just as air bubbles would. Its existence seems almost necessary for the development of overhanging ends, for there are glaciers in the same region which end like ordinary alpine glaciers;[1] and Professor Salisbury tells me that, so far as his observation went, all those, and only those, containing much débris in their lower layers, have vertical or overhanging ends well developed. Professor Chamberlin does not think this is universal.

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[1] This JOURNAL, Vol. III, 823–832. The proof depends on the quasi-viscous theory of glacial movement.

[2] Studies of Muir glacier, Nat. Geog. Mag., IV, 47.

Both of these endings are forms of stable equilibrium; for an increase of velocity, accompanied, as it must be, with an increase of differential velocity, would cause the upper layers to project more, resulting in more breakage; and the surface slope of the lower layers would overhang more, thus presenting a larger surface to the salt water or to the air, as the case may be, and increasing the melting; for we must remember that even in northern Greenland some of the melting comes from contact with the air, which, during the summer, is well above the freezing point.

Variations of glaciers.—Although the sloping surface of alpine glaciers is a surface of equilibrium, it is unstable, and any cause, such as a few years of greater melting or accumulation, which would alter this slope, would destroy the equilibrium, and the surface would tend to depart more and more from its equilibrium form. When the equilibrium of the surface is destroyed by diminished melting or increased flow, the upper layers everywhere advance over the under ones; of course, the higher and more rapidly moving layers advance at a greater rate, but as they are near the névé-line, where the direction of motion makes but a small angle with the glacier's surface, they cause but a very small swelling; nearer the end, however, the direction of motion, making a higher angle with the surface, tends more directly to increase the thickness. This increases the velocity, causing a further departure from the equilibrium surface, and the glacier grows in length with a small increase of thickness in the reservoir, and with a great increase of thickness at its lower end. The ice is rapidly carried off until the upper part of the dissipator becoming thinner, there is a gradual diminution in velocity, and the end of the dissipator having advanced to warmer regions the melting is faster and the glacier begins to recede. This makes it evident that under certain conditions, such as a bed of increasing slope, a glacier, when once the equilibrium between its flow and melting is destroyed, might advance with strides entirely out of proportion to its usual motion, as has frequently happened with the Vernagt glacier.

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[1] CHAMBERLIN, Bull. Geol. Soc. Amer., 1895, VI, 202. SALISBURY, this JOURNAL, 1895, III, 887–890.

A glacier having the equilibrium form corresponding to the climatic conditions prevailing at the time, would lose this form and respond immediately, by a change in length, to the slightest climatic change; whereas, a glacier widely removed from its equilibrium form could not respond by a change in length until the effects of the climatic change had accumulated sufficiently to bring the glacier back to, and carry it beyond, its equilibrium form. Two glaciers, in general similar, but differing in their exposure or slope, might be very differently removed from their respective equilibrium forms, and would therefore respond at different times to a given climatic change; indeed, one of them might be so far removed from its equilibrium form that a climatic change lasting for several years might not be long enough to reverse the condition of retreat or advance in which it happened to be.

Smaller glaciers in general respond more quickly to climatic variations, and we can see why this should be true; for, a change in the amount of snow-fall or of melting, would produce the same change in the actual thickness of two glaciers of different sizes, but subject otherwise to the same conditions. The smaller glacier would, however, experience a greater relative change in thickness, resulting in a greater relative change in flow; it would therefore respond more quickly than the larger glacier; this is confirmed by observation. We should also expect a relatively greater change in size of its dissipator; although, from lack of data, I cannot say that this inference is borne out, still we know that many small glaciers entirely disappear after a few dry hot seasons, and reappear after a few cold wet ones.

This theory offers a connecting link between the theories of Professor Forel and Professor Richter.[1] According to it a glacier will respond to any climatic change, not by the progression of a wave along its surface, but by a change over the whole surface; but this change will not show itself by reversing the phase (the condition of advance or retreat) of the glacier until it has sufficiently accumulated to carry the surface up to and beyond its equilibrium form.

HARRY FIELDING REID.

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[1] See this JOURNAL III, 278-288.